5

CLIMATE VARIABILITY AND TRENDS

Drivers

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> Free lords, cold snow melts with the sun's hot beams
>
> Henry VI, Part II *by William Shakespeare*

5.1 Introduction

Chapter 4 describes the basic components of weather and climate, and a common theme throughout is that there is considerable variability in space and in time. In this chapter, we start by describing and explaining how climate varies by location, by considering the effects of altitude, latitude and other aspects of geography on the climate. We then examine how climate varies over time, starting with differences between night and day, describing the seasons and how they are affected by location, and then describing how and why climate varies from year-to-year and at even longer timescales. Based on what we have learned in Chapters 1–4 the connection of climate to the spatial and temporal risk of infectious diseases, malnutrition or hydro-meteorological disasters can now be made.

5.2 How does climate vary spatially?

The average temperatures and rainfall for January and July across the globe are shown in Figure 5.1 and Figure 5.2. Why are some places hotter, or drier, than others? The most important factors affecting the spatial variation of climate are:

- Altitude
- Latitude
- Contrasts between the effects of land and sea
- Different land surface conditions

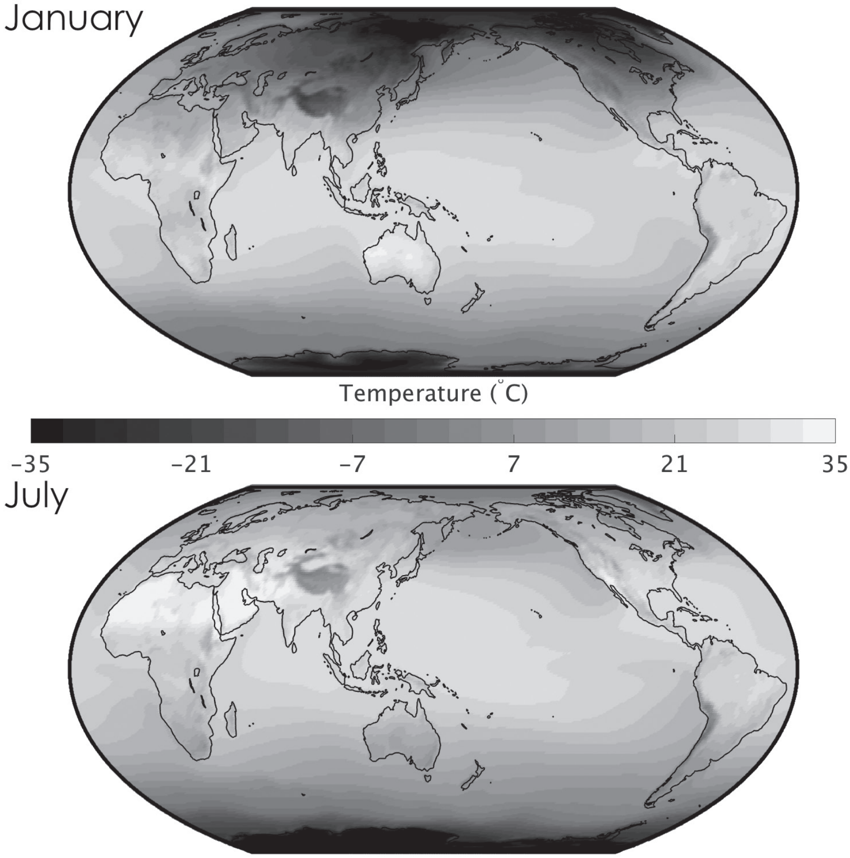

FIGURE 5.1 Average temperatures for January (top) and July (bottom) 1981–2010. *Data source*: ECMWF Interim Reanalysis, for 1981–2010[1]

5.2.1 Climate and Altitude

5.2.1.1 Temperature and altitude

An important lesson of Chapter 4 has been that the air is heated from Earth's surface rather than directly by the sun. This distinction has profound effects on how temperatures vary horizontally and vertically. Consider first, vertical changes in temperature. Even from a cursory inspection of Figure 5.1, an effect of altitude on temperatures is evident from the relatively cold temperatures over the Himalayas, the Andes and other high mountain ranges. It is easier to see the effects of altitude by looking at how temperatures change above a specific location (Figure 5.3; in this case, Brookhaven, NY, but the general pattern is similar in most locations). There are two main features:

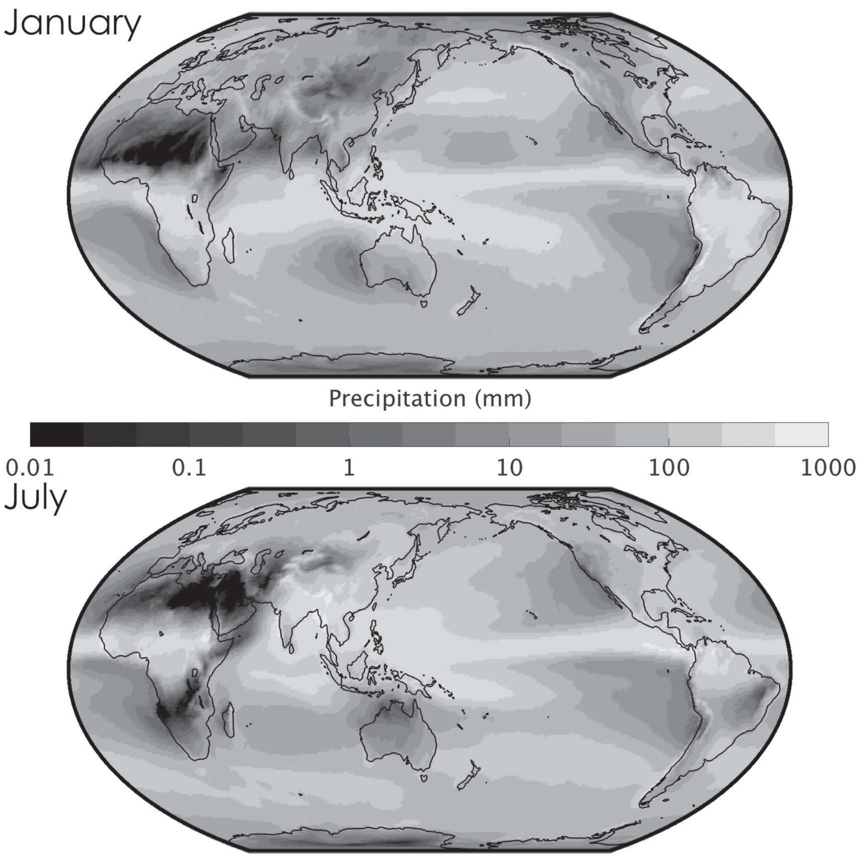

FIGURE 5.2 Average rainfall for January (top) and July (bottom) 1981–2010.
Data source: ECMWF Interim Reanalysis, for 1981–2010[1]

- Temperature decreases rapidly at a near-constant rate for about the first 10 km (jet airplanes fly at about this altitude, where you may have noticed recordings of outside temperatures of around –60 °C or –70 °C).
- Above about 10 km, temperatures decrease more slowly, and start to increase above about 20 km.

The air is warmest near the surface because that is where the air is heated by the Earth, but the secondary peak in temperatures at about 25 km altitude is where the sun can heat the air up directly because of the presence of ozone at these altitudes. Ozone is one of the few gases that directly absorbs the sun's rays – specifically ultraviolet radiation (§ 4.2.6) and so at these high altitudes the sun heats the ozone directly, which, in turn, heats up the gases in the immediate vicinity.

The rapid temperature decrease in the first 10 km above the surface is a direct effect of a decrease in air pressure at higher altitudes (§ 4.2.7). Air pressure decreases

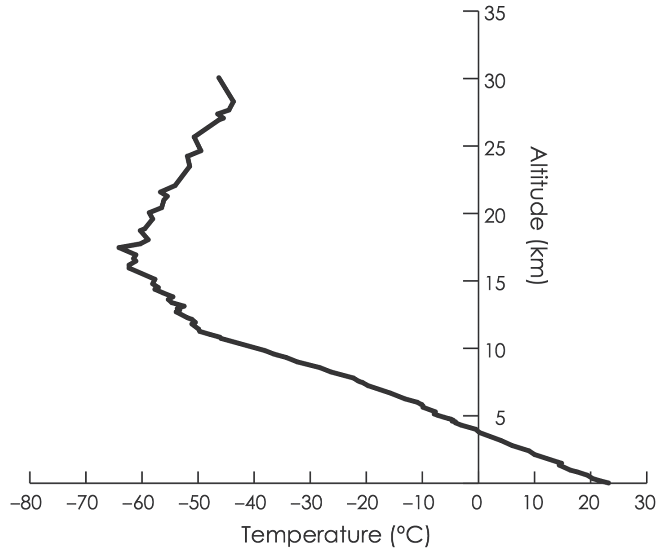

FIGURE 5.3 Temperature as a function of altitude over Brookhaven, NY, at 08h00 local time on 3 August 2017
(*Data source*: https://ruc.noaa.gov/raobs/)

with elevation because there is less weight of air pressing down from above. As incredible as it may sound, if you stand on the beach at the sea there is about a tonne of air pressing down on you. At sea-level the air is squashed because of all that weight, but higher up there is less weight of air above and so the air is less dense and can expand into more space. For an equivalent reason, we do not put our tomatoes in the bottom of the shopping bag, otherwise the heavier items on top will squash them. The amount of air decreases by about 11% for every kilometre in altitude.

Because of this decreasing air pressure, air expands as it rises, while as it descends it compresses; but, by expanding, the air cools, and by compressing it warms up. This cooling and heating effect can be reproduced with a simple experiment. If you blow on your hand, your breath feels cold because it is forced through a narrow hole in your mouth and then expands as it gets into the open air; but huff and your breath feels warm because it is not expanding into the open air. Similarly, if you pump up a bicycle tyre you may notice the pump gets hot because of the air compression, but the air escaping from the valve or through a puncture feels cold.

The cooling of air as it expands when it rises occurs at a predictable rate known as the *lapse rate*. As a reasonable rule of thumb, the temperature decreases about 1 °C every 100 m, which is approximately the rate of cooling you would feel when ascending in a hot air balloon. Conversely, air warms by about 1 °C every 100 m as it sinks. That lapse rate is the same in hot and cold air, and so it is the same in the tropics and the extratropics. However, there is an important complication: it is not the same in humid and dry air. Air only cools at this rate when it is unsaturated (i.e., the relative humidity is less than 100%).

The heat that was originally used to evaporate water is released if the air is cooled enough to condense the vapour back to water (see the discussion on latent heat in § 4.2.1). This latent heat partly reheats the air so that the air cools more slowly as it ascends once it becomes saturated. On average, the decrease is about 0.6 °C every 100 m, but the actual rate depends heavily on how much water vapour does condense, which in turn depends on the temperature. The amount of water vapour required to saturate air increases exponentially with temperature, and so hot humid air contains a lot more latent heat than cool humid air (§ 4.2.3). Therefore, as hot humid air cools, it condenses large amounts of water vapour and releases large amounts of latent heat, but cold humid air releases only small amounts of latent heat. As a result, the lapse rate is much lower in hot humid air than in cool humid air. One important effect of this lower lapse rate is that rising hot humid air is likely to stay hotter than its surroundings because it cools only a small amount with height, and therefore it will remain buoyant. The buoyancy of hot humid air because of the condensation of large amounts of water vapour is critical in forming heavy rainfall and violent storms in the tropics. It is one reason why hurricanes and typhoons, for example, form only in the tropics (§ 4.2.8).

The lapse rate indicates how quickly air cools as it rises, but that does not mean that temperatures decrease by that amount if you climb up a hill. If a wind blows up a hill, the air will cool as it ascends, but because the air is still near the surface, that cooling may be offset by heating from the hill itself. The actual decrease in air temperature uphill is affected by many factors including the orientation of the slope and exposure, the weather and the time of day. It may even be warmer up the hill if the air is clearer there or if the slope faces the sun more directly. It is important to be aware of such micro-climatic variations when temperature thresholds are being used to guide decision-making processes. Nevertheless, temperatures do generally decrease uphill. In the Himalayas and the Alps, for example, temperatures decrease uphill at a rate of about 0.4 °C to 0.7 °C every 100 m. An average rate of about 0.6 °C (about the same as the average lapse rate in moist air) is a reasonable rule of thumb. Understanding just how temperatures cool at higher altitudes is important for malaria planning (see Case Study 5.1).

The rapid decrease in temperatures in the first 10 km that is shown in Figure 5.3 is typical of most places and times of year. However, under certain weather conditions temperatures can increase for the first few hundred metres. Such an increase is known as an *inversion*.

Inversions can occur when warm air blows over a colder surface. Fog may form from this process if the cooling is enough to produce condensation. This type of fog is common in mid-latitude maritime climates (see Box 4.1 for definitions of climate regions) during the winter. It also occurs in the subtropics along the west coast of many of the large desert regions, where winds may be cooled by a cold sea surface.

CASE STUDY 5.1 ELEVATION USED IN PLANNING MALARIA CONTROL PROGRAMMES

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The East African highlands have long been considered desirable for human habitation because of their rich soils and low levels of infectious diseases such as malaria. Since air temperature decreases with elevation (§ 5.2.1.1), the degree days required for malaria parasite development become increasingly difficult to achieve the higher up the mountain one goes until a minimum temperature threshold is reached where transmission is no longer possible. Much of the Ethiopian highlands are above the altitude at which this threshold is reached, with low temperatures forming a natural barrier to malaria transmission. As such, elevation is used in malaria control planning by the Ethiopian Ministry of Health[2] as high altitudes (> 2000 m) are considered malaria-free. Below 2000 m and outside desert areas, the country is considered endemic for malaria and therefore included in routine control efforts. However, the association of temperature and altitude is not uniform across the country: there is a wide range of temperatures for stations at similar altitudes from different parts of the country (see Figure 5.4), while the overall relationship is close to the

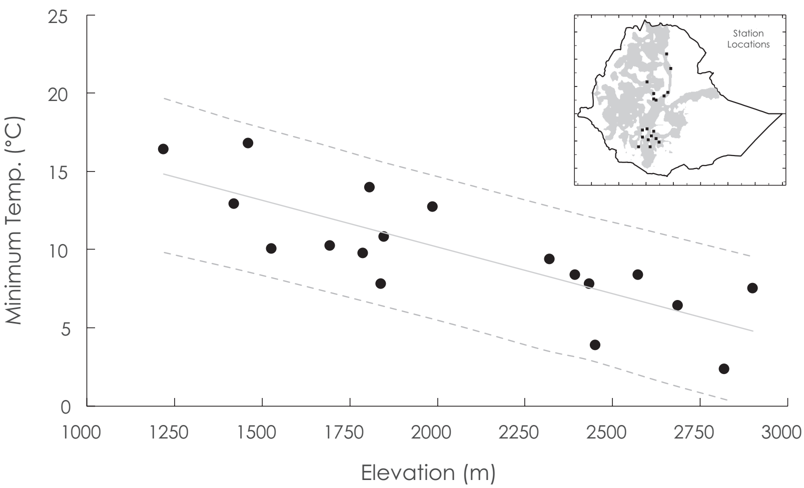

FIGURE 5.4 Climatological (1981–2010) monthly average minimum temperature (°C) as a function of elevation for 18 stations in the Ethiopian Highlands (locations shown in insert). Points indicate the lowest climatological monthly minimum temperature during the calendar year, with the solid line showing a least squares linear fit and dashed lines indicating the 95% confidence limits for predicted values

expected lapse rate (§ 5.2.1.1). A new malaria transmission boundary map was created using high resolution gridded temperature data from the National Meteorological Agency.[3] This boundary map revealed that there can be a 1000 m difference in elevation for the same minimum temperature threshold depending on where it was measured (e.g., which side of the mountain).

Some of the strongest inversions occur in the extratropics on cloudless winter nights. Such weather conditions are typical of periods of high air pressure (§ 4.2.7) where the air is descending and therefore warming. As it approaches the ground, this descending air may become hotter than the Earth's surface, especially on a clear winter night when the ground can cool down quickly. In such weather conditions, air pollution may become severe, partly because more fuel is burned for heating, but also because the inversion prevents the pollution from dispersing easily. Because the air aloft is warmer, the pollution cannot easily rise to higher altitudes where it could be dispersed by stronger winds. This cause of severe air pollution episodes is often a problem in cities that have a winter dry-season and weak winds, such as New Delhi, Santiago, Mexico City and Ulaanbaatar. Such weather conditions are also responsible for some of the most severe episodes of pollution in cities with more variable weather, such as London's Great Smog of 1952, and New York City's 1966 smog. In 2017 New Delhi had the unenviable title of the most polluted city on the planet. On 8 November 2017, the city's air quality index measured in the range of 700–1000; the US Environmental Protection Agency considers anything over 300 to be hazardous. The extreme smog resulted from smoke caused by burning of stubble on nearby farms coinciding with a temperature inversion that kept the smoke and other polluted air in the city for several days.

5.2.1.2 Humidity and altitude

Just as temperatures decrease with altitude, so does humidity. How humidity decreases depends, in part, on how it is measured (see Box 4.3), but it is sufficient to note that the amount of water vapour in the air decreases with altitude largely because of the effects of decreased temperature and air pressure. The cooler temperatures and lower humidities mean that heat stress (see Box 4.2) is unlikely to be a major health problem at high altitudes, both now and in the future.

5.2.1.3 Wind and altitude

Although heat stress at high altitudes is unlikely to be a problem, wind chill is a bigger risk, not only because of the colder temperatures, but also because wind speeds tend to increase with altitude. Wind speeds increase with altitude partly because there is less friction with the surface, and partly because air pressure gradients generally strengthen. Wind chill combined with extremely low temperatures can be exceptionally hazardous for mountaineers.[4]

5.2.1.4 Rainfall and altitude

The complicated relationship between humidity and altitude is reflected in that between rainfall and altitude. Nevertheless, some general patterns can be identified. Up to a point (discussed further below), rainfall increases at higher altitudes on the windward (and often wetter) side of mountains, and so knowledge of the prevailing wind direction (§ 4.2.4) is important for understanding mountain climates. The increase in rainfall with altitude is a result of the orographic effect (§ 4.2.2): winds blowing towards a mountain are forced to rise, resulting in cooled air, from which water vapour may condense to form clouds and possibly rainfall. On the leeward side, the winds descend the mountain, the air warms, clouds are evaporated, and so rainfall is inhibited.

The contrast between the relatively wet windward and dry leeward sides of mountains is evident across the globe. However, the rate of change in rainfall with altitude on the windward side is complicated. On the one hand, the higher up the mountain, the cooler the air becomes, and so the greater the likelihood that water vapour will condense. On the other hand, much of the water vapour at high altitudes may already have condensed and so the tops of mountains may be above the clouds. Even if there is still moisture left, the low temperatures at these high altitudes mean that only small amounts of condensation will occur because of the exponential relationship between temperature and saturation that was discussed in §§ 4.2.3 and 5.2.1.2. Therefore, rainfall starts to decrease with altitude beyond an 'elevation of maximum precipitation'. For most of the globe this elevation is somewhere between about 1 and 2 km, but in polar latitudes, the air is too cold to hold much moisture. Here rainfall (or, more typically, snowfall) is at a maximum at the base of a mountain, and decreases with elevation. In the humid tropics, the elevation of maximum precipitation tends to be on the lower side of the 1–2 km range because of the large volumes of condensation that occur at low altitudes where the air is most humid.

If nothing else is clear beyond the fact that this relationship between rainfall and altitude is complicated, you have understood the main conclusion! What generalizations, if any, can be drawn? The altitude of maximum rainfall tends to be higher in the mid-latitudes than in the tropics and high latitudes, and for similarly complicated reasons, the altitude of maximum rainfall tends to be relatively high in dry areas, in the dry season, and in the warm season.

5.2.2 Climate and latitude

5.2.2.1 Rainfall and latitude

As just noted, not much snow falls near the Poles because the air is too cold to hold much moisture. For this reason, Antarctica (the left side of Figure 5.5) is the driest continent on Earth. Antarctica is drier than the Arctic (the right side) because the South Pole is so much colder than the North (Figure 5.6). However, rainfall does

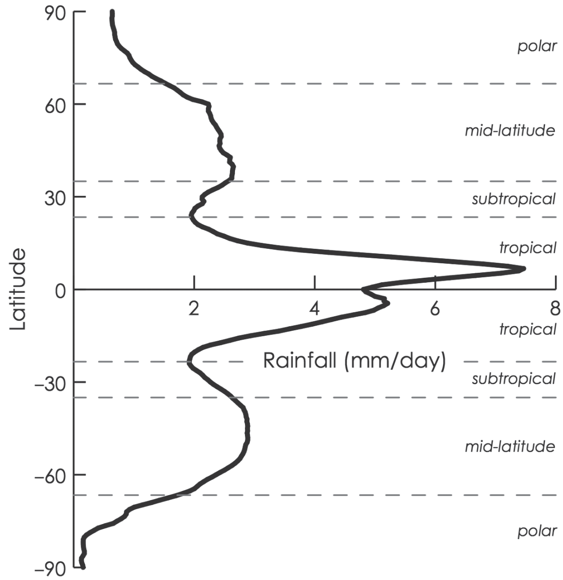

FIGURE 5.5 Rainfall (and snow) as a function of latitude (the South Pole is at the top; the North Pole at the bottom).
Data source: ECMWF Interim Reanalysis, for 1981–2010[1]

not decrease simply from the warm tropics to the dry polar latitudes (Figure 5.5): there is a band of deserts through the subtropics as well (including the Sahara in the Northern Hemisphere, and the Namib and the deserts of Australia in the Southern), which are visible in Figure 5.2. This arid belt at about 30° latitude occurs in both hemispheres, and is caused by poleward moving air from the equator sinking in the subtropics as a result of Earth's rotational effects.

Most rainfall occurs near the equator where solar heating, and therefore evaporation of surface water, are at a maximum, and where high temperature and humidity can encourage the formation of heavy rainstorms (§ 4.2.8). There is a slight local rainfall minimum on the equator itself because of a surprisingly cold sea surface in the eastern Pacific Ocean (Figure 5.1) which is relatively dry as a result (Figure 5.2). This pattern of a cold and dry eastern equatorial Pacific is occasionally disrupted as part of El Niño (Box 5.1).

5.2.2.2 Temperature and latitude

The South Pole is colder than the North Pole (Figure 5.6) primarily because of differences in altitude (§ 5.2.4). The North Pole is near sea-level whereas much of Antarctica is above 2 km; Antarctica's altitude contributes to at least a 15 °C cooling. The temperature difference is exacerbated by land–sea contrasts: the South Pole is straddled by the continent of Antarctica while the North Pole is straddled by the

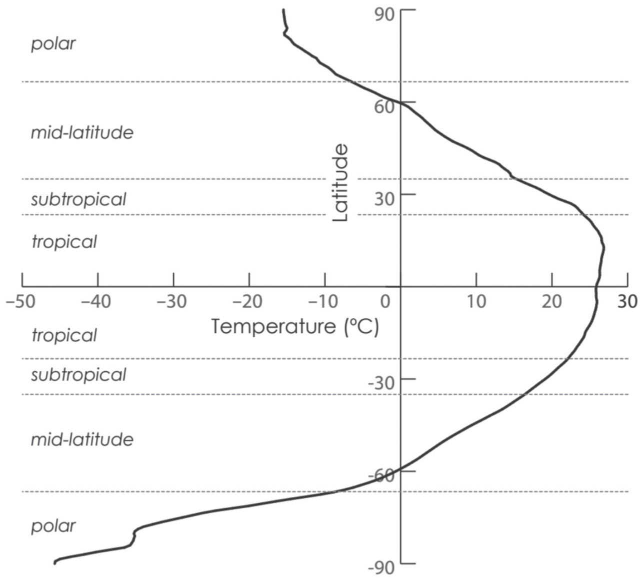

FIGURE 5.6 Temperature as a function of latitude (the South Pole is at the top; the North Pole at the bottom).
Data source: ECMWF Interim Reanalysis, for 1981–2010[1]

Arctic Ocean (see § 5.2.6 on the contrasting effects of land and sea). Areas near the equator have the highest temperatures on average through the year, and temperatures are consistently high through the year. However, the most extreme temperatures (well above 50 °C) occur away from the sea in the subtropics (specifically, in Kuwait, Libya and California), where land-surface temperatures exceeding 80 °C have been recorded. These locations have a large annual range in temperature: they experience relatively cold winter and night-time temperatures compared to at the equator (§§ 5.3.1 and 5.3.4).

Why do air temperatures decrease towards the Poles? The answer has to do with the sun's heating of Earth's surface. The air is heated by Earth's surface, but the surface temperature depends largely on the intensity and duration of the sun's heating. During the winter, because the sun does not appear at the Poles at all, it does not heat the surface for weeks or months (see § 5.3.3). During the summer, although the sun is up all day, it remains low in the sky, which has two effects:

- The intensity of the sun's heating is weak. The sun's heating is most intense when the sun is high in the sky – directly overhead. If you shine a torch directly on the ground it will make a small bright circle, but shine it on the ground at an angle and it will make a larger and dimmer oval. The torch is emitting the same amount of light, but that light spreads over a larger area simply because of the angle. The same is true of sunlight: at about 60° latitude, sunlight is spread over about twice the area it as at the equator because Earth is spherical. Near the Poles (or at any latitude in the morning, or in the mid-latitude winter), the sun is low in the sky and so it cannot heat the surface very intensely, but near the equator (or at high noon in summer) the sun is more directly overhead and its heating is more intense. Averaged over a year, the Poles receive the least amount of the sun's energy, but for a short time in the summer they do receive the maximum amount per day. Despite the low height of the sun, the Poles get more sunlight per day in their peak summer than the equator gets at any time of year. The 24-hour summer days at the Poles more than offset the low height of the sun in the sky.

- Much of the sunlight at the Poles is reflected and so does not heat the ground at all. Despite the 24-hour summer daylight in the summer much of that sunlight is reflected, and so it does not heat the surface at all (see the discussion on albedo in § 4.2.7). Therefore, the Poles remain cold year-round. The highest recorded temperature at the South Pole, for example, is well below freezing point – about –12 °C.

5.2.3 The effects of land and sea

It takes an enormous amount of energy to heat up water: about four times more than for the air, and even more than for the ground. On a sunny summer's day, you can burn your feet on the paving stones around a swimming pool, but the water stays a pleasant temperature. Occasionally there are news reports of people frying eggs on a car bonnet in summer, but never about boiling an egg in a bucket of water left in the sun. It is why water is used as a coolant in car radiators, for example – water will cool down most things (metals, air, etc.) very effectively, while heating up only a small amount itself in the process.

5.2.3.1 Effects on temperature and the seasons

Because it takes so much energy to heat water, the oceans and large lakes have a major dampening effect on climate, and even rivers and small lakes may have a detectable effect. The difference in temperature between the warmest and the coldest times of the year (Figure 5.7) is one measure of the dampening effect – where the difference is small the seasons are less extreme. This annual range in temperature exceeds 20 °C over large parts of North America and Eurasia, whereas over much of the oceans the annual range is less than 10 °C. The contrast in the annual range between land and sea is strongest in high latitudes where differences in the amount of sunlight between summer and winter are extreme because of large changes in the length of the day (§ 5.3.4). However, there is also an apparent longitudinal effect: in Eurasia and North America the western sides of the continents have much smaller ranges (i.e., milder seasons) than do the eastern sides. This east–west contrast is a result of the prevailing wind direction (§ 4.2.4), which is from the west at these latitudes. Westerly winds bring air from over the sea to the western sides of the continents, but from over the

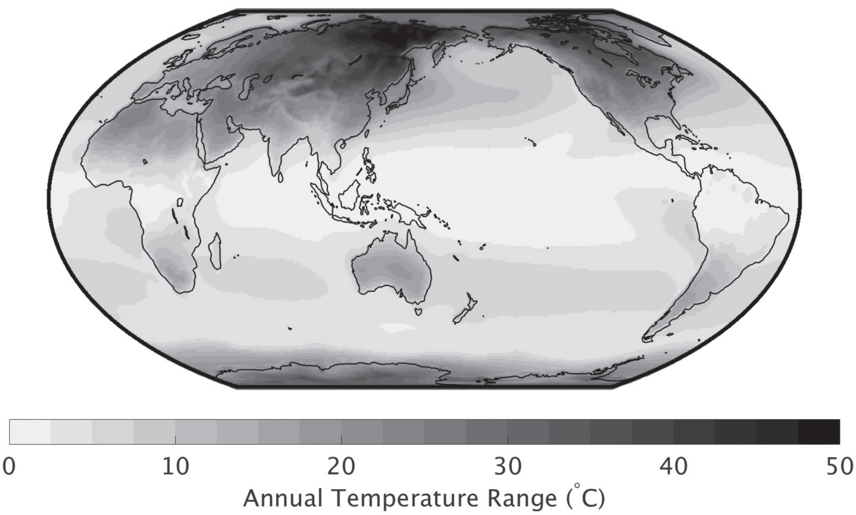

FIGURE 5.7 The annual range in temperature, calculated as the difference between the warmest and coldest mean monthly temperature. (Scale is from 0 [black] to 40 [white].)
Data source: ECMWF Interim Reanalysis, for 1981–2010[1]

land to the eastern parts. Therefore, the apparent longitudinal effect is actually an example of the effects of the distribution of land and sea.

As well as dampening the seasons, the oceans also delay them. Water heats up so slowly, that the warmest time of year at coastal areas is delayed until well after the time of strongest solar heating. Similarly, the sea cools down very slowly – once water is heated up, it stays hot for a long time – and so the coldest months in coastal areas occur well after the shortest day of the year. This delaying effect on the seasons is most evident in the mid-latitudes, where sea temperatures do not reach a maximum until about two months after the summer solstice.

5.2.3.2 Effects on humidity and rainfall

Because the sea and major lakes are the primary sources of moisture in the atmosphere, humidity is generally higher in coastal areas than it is inland. However, the influence of the oceans on the distribution of humidity and rainfall is much more a reflection of sea-surface temperatures than it is of distance to the coast. Where the sea is cold, rainfall and humidity are likely to be low, but near warm oceans humidity is likely to be high and abundant rainfall may occur. This effect of sea-surface temperatures on rainfall is most evident in the subtropics (at around 30° latitude; see Figure 5.1): here the oceans are much colder on their eastern side than on their western side because of how ocean currents are affected by temperatures and Earth's rotation. The amount of water vapour that air can hold increases exponentially with temperature, which means that over warm oceans the air can hold much more moisture than over cold oceans. As a result, the western side of continents (i.e., bordering the colder eastern ocean) tend to be much drier than the eastern side (bordering the warmer western side of the ocean).

Land–sea contrasts are important in the formation of the monsoons (see Figure 4.3). In the summertime, the land heats up more quickly than the sea and so the hotter air over the land becomes less dense. Winds will blow inland from the cooler sea to compensate for the difference in density. In winter, the land cools down more than the sea, and so the colder air over the land becomes denser and will blow towards the sea. There is therefore a strong seasonal contrast in the prevailing winds, and typically in the rainfall too: the winds from the sea are likely to bring humid rainy weather, but those from the land bring dry conditions. The most pronounced example of this contrast is the Indian Monsoon.

Similar mechanisms can apply on a smaller-scale with major lakes to affect local climate. In temperate zones, lakes moderate the temperatures of the surrounding land, cooling the summers and warming the winters, such as around the Great Lakes of North America. Lakes also act like giant humidifiers, increasing the moisture content of the air. In the winter, this moisture contributes to heavy snowfall, known as 'lake effect' snow. Large lakes can also develop their own storm systems. For instance, Lake Victoria is prone to deadly night-time storms that kill thousands of fishermen annually. Forecasting lake storms hours or days in advance has the potential to save many of these lives.

5.2.4 The effects of land-surface type

5.2.4.1 Urban heat islands

Although the contrast between land and sea is the most important effect of Earth's surface on climate around the globe, other contrasts are also important. The importance of snow and ice formation is discussed in §§ 4.2.7 and 5.2.2. Another contrast that has important local effects is that between urban and rural areas. In large urban areas (with populations of about one million or more), the construction materials, the shapes of buildings and the impacts of human behaviour can increase the temperature by around 1 to 3 °C on average. The areas impacted by these increases in temperature are called urban heat islands. The heat island usually is strongest towards the centre of the city, but local land use patterns modulate the heat island significantly, and so temperatures can differ by a few degrees within only a few blocks. The heating effect is strongest at night and in the cold season (and is strengthened further if any snow is removed manually) when warming of as much as 12 °C has been measured, and significantly contributes to heat wave risk during the hot season (Case Study 7.2). The heat island combined with various air pollutants can also increase rainfall over, and downwind of, cities. (The pollutants increase rainfall because water vapour condenses more easily into large drops – heavy enough to fall as rain rather than just form clouds – around particles.) The effect on rainfall is strongest in the hot season.

5.2.4.2 Deforestation

There is frequent mention of the contribution of deforestation to climate change. Deforestation does affect global climate primarily because of the release of large amounts of carbon into the atmosphere from the trees and the soil, rather than through large-scale impacts of changes in the Earth's surface. It can also have an important effect on local hydrology – deforestation can contribute to increased flooding – but the impacts on climate are often exaggerated.

Although deforestation tends to impact local hydrology more than local climate, some local climate impacts can occur. Small-scale deforestation often results in an increase in local rainfall because of increased surface heating – in a similar way to the urban heat island effect on rainfall. Large-scale deforestation, on the other hand, can result in decreased rainfall because there is less moisture available to evaporate from the soil and vegetation.[5] A decrease in evaporation from the land may be more important than one might first guess: about 40% of rainfall that falls over the land is evaporated from the land rather than the sea, and that proportion reaches 70% in parts of the Amazon forest. A large amount of the rainfall over the land is therefore 'recycled' – it previously fell as rain and has not yet returned to the sea.

The rainfall decrease after large-scale deforestation, combined with the reduction in moisture in the soil and vegetation canopy, have important effects on surface temperature. Less energy is needed to heat and evaporate water, and so the ground can be heated up more quickly even though the removal of the dark vegetation cover means that more sunlight is reflected. The increased surface temperature raises the air temperature, and so large-scale deforestation can result in a hotter and drier local climate.

5.2.5 Climate and spatial scale: How big is a heat wave, or a drought?

In the previous sub-sections we have seen how climate varies over large distances – at different latitudes, with proximity to the coast, or on the different sides of a mountain – but how does climate vary over shorter distances? Is the next farm or town experiencing similar weather and climate conditions as are occurring here? The West African story *Anansi and the chameleon*[6] correctly indicates that it is quite possible for the next farm to remain dry even when it is raining heavily over your farm. However, the situation may not be as uncertain for sunny weather: if your farm is experiencing drought, it is quite likely that the next farm is experiencing the same drought. Drought is a climate condition (one day of sunny weather does not constitute a drought), and similar climate conditions occur over larger areas than do similar weather conditions.

5.2.5.1 Spatial scales of temperature

Although temperatures can vary by many degrees in short distances (because of terrain or because of marked land-use differences in cities, for example), temperature anomalies (how unusually hot or cold it is) tend to be similar over large areas, extending to about 1500 km almost regardless of the timescale of interest. Any small-scale differences in temperature anomalies are quickly smoothed over by changes in wind direction because air density, and hence winds, are strongly affected by temperatures (see § 4.2.4).

In the extratropics, temperatures are strongly affected by the large-scale wind direction because climatological temperature differences are large over distances of only a few hundred kilometres (see Figure 5.1, and § 5.2.5). Winds from a relatively warm location will increase the temperature, but the temperature will fall if the wind comes from a cold location. If the winds are northerly or southerly, the interpretation is almost always simple: winds from the poleward latitudes are likely to be cold, those from more tropical latitudes are likely to be warm. An example over the USA is shown in Figure 5.8, where northerly winds bring cold air to the central states but southerly winds bring warm air to the eastern states. For easterly and westerly winds, the effect is more complex: if the new wind is coming from a continental interior it is likely to be unusually hot during the summer, but cold

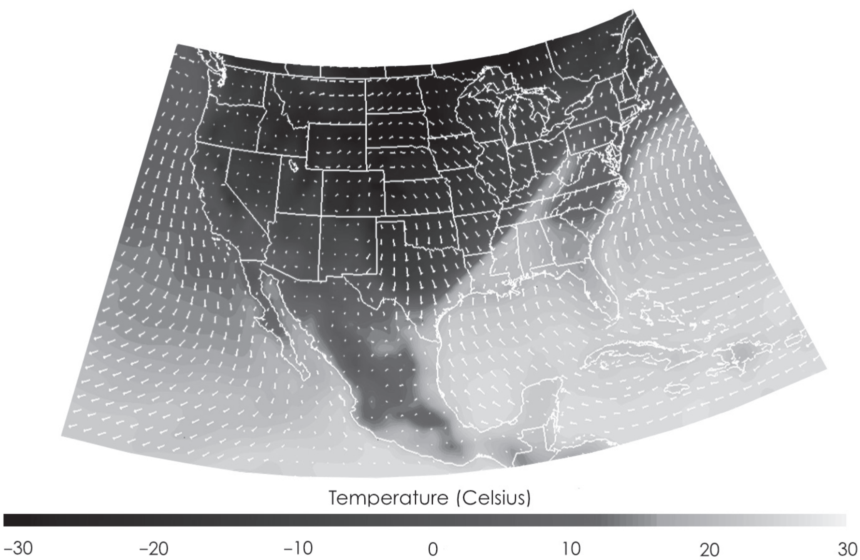

FIGURE 5.8 Example of a cold front, occurring on 26 November 2015 over part of the USA. The front lies along the boundary between the warm air to the southeast (light shading), and the cold air to the northwest (dark shading). The 2 m wind speeds and directions are shown by the arrows.
Data source: ECMWF Interim Reanalysis, for 1981–2010[1]

during the winter, and the opposite is the case if the wind is coming from the ocean. These effects of changes in the prevailing wind on mid-latitude temperatures apply day-to-day, and also year-to-year, and are inherently large-scale; in contrast, any highly localized or short-lived change in wind direction is likely to have only a negligible effect on temperature.

Warm and cold spells are likely to extend over large areas, and so if it is unusually cold in one town it is likely to be unusually cold in the next town also. However, in the extratropics at weather timescales there is often a narrow band (called a front) where the temperature changes markedly over a short distance (Figure 5.7), most commonly between autumn and spring. These fronts typically move at about 30 to 80 $km.h^{-1}$, and so they are not usually evident when data are averaged over more than a few days.

In the tropics, changes in wind direction do not have as strong an effect on temperature as they do in the extratropics, and so temperatures do not vary as much from day-to-day or year-to-year (§ 5.3). An exception is an occasional large-scale incursion of extratropical air in the cold season (if there is a cold season). Tropical temperature anomalies reflect changes in surface heating because of changes in cloudiness, and/or because of changes in ocean temperatures and currents, as in the case of El Niño (Box 5.1). The cloudiness changes can be highly localized, and on small mountainous islands temperature anomalies can differ considerably on opposite sides of the island. However, in most cases, the heating anomalies are large-scale, and so temperature anomalies are correlated over hundreds of kilometres.

5.2.5.2 Spatial scales of rainfall

Rainfall anomalies are less spatially coherent than those for temperature, and there is a strong dependence on timescale. The spatial coherence depends on the type of rainfall (§ 4.2.2). Convective rainfall is highly localized at weather timescales, such that individual rainfall events are hard to capture with fewer than two rain gauges per km^2.[7] Convective rainfall is common in the mid-latitudes in summer, and year-round in the tropics. One may not need to drive very far (or wait very long) for the rain to stop. However, convective showers do generally track the larger scale wind patterns and so a series of convective showers over a few hours or days is likely to deposit rainfall over a reasonably large contiguous area, much larger than the area rained upon at any moment. If it is not raining on the neighbour's farm now, it may well do so shortly, perhaps after it has stopped raining on yours, or perhaps in the next few days.

As its name suggests, large-scale rainfall is more spatially coherent than convective rain. However, even within large-scale cyclones (§ 4.2.8) rainfall intensity can vary considerably over very short distances. While it may be raining on your farm and your neighbour's farm, one of you is likely receiving quite a bit more rain than the other.

Although extratropical cyclones can be thousands of kilometres in diameter, the weather is very different in different parts of the cyclone. Cyclones are not a large-scale organization of one type of weather; instead, they are a large-scale organization of air circulation involving different types of weather. Each cyclone is unique, but as a simple generalization the following types of weather are typical. Ahead of the warm front (Figure 4.6), there is usually continuous cloud cover, and possibly light to moderate continuous rain or snow. Behind the warm front, the air feels more humid, and there may be patchy clouds. The cold front often brings heavier rain or snow with thicker clouds than in the warm front.

What can we conclude about the spatial coherence of rainfall? The amount of rainfall is highly localized at most timescales, and correlations beyond about 100 km occur only at seasonal scales. For rainfall occurrence (at the daily timescale) or frequency (for longer timescales), spatial coherency is stronger, but is usually limited to about 450 km. This is a sobering thought given that it is not uncommon in analyses of rainfall and health outcomes to use meteorological data from stations tens or even hundreds of kilometres from the site of origin for the health data.

5.3 How does climate vary temporally?

Weather and climate can vary considerably at almost all timescales, but the amount they do vary differs considerably from place-to-place, and at different times of the year. In the following sections, the effects of the time of day and the season are described and explained. Reasons why the climate is not the same every year are considered in § 5.4.

5.3.1 How does the time of day affect the climate?

5.3.1.1 Temperature

Perhaps the most immediately noticeable difference in climate conditions is the contrast in temperature between night and day. Variations through a day are called diurnal variability. The minimum temperature most frequently occurs at about sunrise (Figure 4.1), but even though the sun's heating is strongest at high noon (ignoring any complications from cloudiness), the hottest time of day may not occur until mid- or even late-afternoon because it takes time to heat up Earth's surface. The cycle of heating and cooling is not perfectly symmetrical, and so the mean temperature is generally slightly less than the average of the minimum and maximum.[8]

In humid conditions, perhaps near the coast or when the air is unusually humid, the maximum temperature occurs later than it does in more arid conditions because of the slowness of water to heat up (§ 5.2.3). In the tropics, the change in temperature during the day almost always follows a similar pattern to that shown in Figure 4.1. However, in the extratropics, the change in temperature can be strongly affected by changes in wind direction (perhaps associated with the passing of a cold front similar to that shown in Figure 5.8; § 5.2.5), and so it is possible for temperatures to decrease during the morning, or increase during the night. Nevertheless, the average diurnal cycle in the extratropics is similar to Figure 4.1. The diurnal

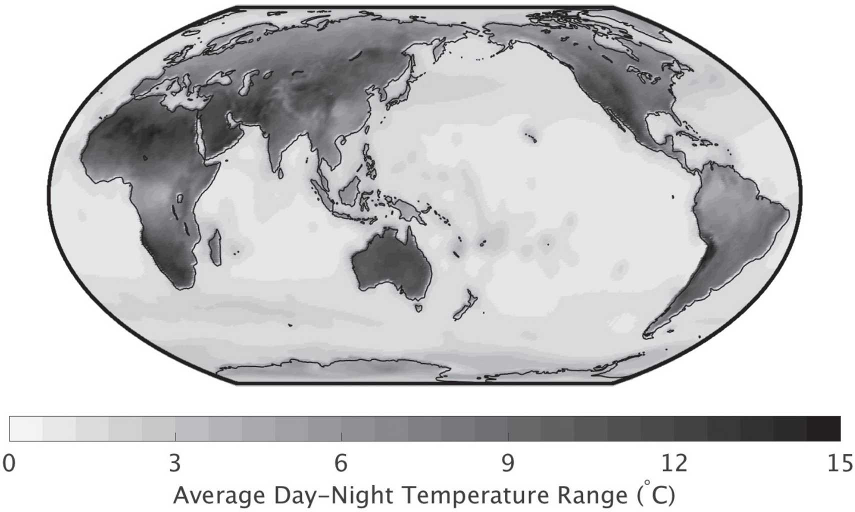

FIGURE 5.9 Average daily-range in temperature for 1981–2010.
Data source: ECMWF Interim Reanalysis, for 1981–2010[1]

range (Figure 5.9) varies considerably depending on humidity, including distance from the coast, and shows some similarities to the annual range (Figure 5.7). The effect of land–sea differences is much stronger for the diurnal range than it is for the seasonal range in temperatures because of the dominant effect of latitude on the annual range.

5.3.1.2 Rainfall

The diurnal variability in rainfall is more complicated than that for temperature, and depends on location (including altitude), time of year and rainfall intensity. Thus, it is possible to provide only some simple generalizations here. Because convective rainfall is dependent on heating and cooling of the air, it has a stronger dependence on the time of day than does large-scale rainfall. Convective rainfall peaks late-afternoon, at about the time of maximum temperature, but there is also a secondary peak a little after mid-night when cooling at night releases latent heat. Large-scale rainfall, on the other hand, does not have a strong diurnal cycle. Orographic rainfall often represents an enhancement of convective and / or large-scale rainfall so diurnal differences are similar to those just described. However, the diurnal cycle of rainfall is made more complicated in mountainous areas.

5.3.1.3 Winds

The time of day can have an important effect on local wind patterns. Wind turbulence is strongest during the daytime because of surface heating. Differences in heating between land and water bodies can create local land and sea/lake-breeze effects. These breezes are similar to the monsoons, but they occur on a daily timescale and at a smaller scale; they do not dominate the large-scale wind patterns in the way the monsoons do. Another example of a diurnal wind reversal can occur in mountainous regions. Air that is near a mountain is close to the Earth's surface – the source of heating – and so this air heats up and cools down faster than air at the same altitude further away from the mountain. During the day, the hotter air near the mountain may rise, creating a warm, uphill wind (and possibly causing rainfall), whereas at night the air near the mountain cools down quickly, resulting in a cold, downhill wind. Mountain valleys can therefore become very cold at night-time.

5.3.2 How long do weather patterns last?

Asking how long weather patterns last is like asking: how long is a piece of string? There is no way of providing a meaningful general answer to the question, and it can only be addressed in the specific: how long will *this* weather pattern last? Rainstorms, for example, move at a wide range of speeds or can stay stationary, and so can last for seconds to days or anything in between. Notwithstanding, a few generalizations are possible. For example, there is a relationship between spatial and temporal scale: convective rainfall, as well as being more localized than large-scale rainfall, is often more short-lived.

If large-scale weather conditions can have a significant impact on Earth's surface then they can have a noticeable effect on subsequent weather. For example, if snow settles then subsequent sunlight will be reflected rather than absorbed by the surface and so temperatures will probably decrease. If snow occurs over very large areas (10,000s km^2) weather patterns may be affected for the next few weeks. Similarly, after a period of dry conditions during the summer, the dry soil may be able to heat up to extremely high temperatures, drying the soil even further. The ground can then get even hotter because there is less energy used for evaporation of any remaining soil moisture. Therefore, the drying of the land surface is often an important precursor of heat waves. Dry air can inhibit the development of rain clouds thus allowing hot dry conditions to persist. However, the higher temperatures may also help to generate a sufficiently strong sea breeze to bring more moisture; so there is no simple relationship.

In the extratropics, persistent weather conditions are most frequently associated with 'blocking' episodes, so called because the usual eastward migration of weather patterns is inhibited by stationary weather systems. These blocking patterns are linked to persistent patterns in the jet streams, which are a major control on the development and movement of weather systems in the mid-latitudes (§ 4.2.8). Blocking can last for days to weeks. Because of some rather complicated effects of land and sea contrasts, blocking occurs most frequently in the Northern Hemisphere, and is most frequent in spring and over the eastern Pacific and Atlantic Oceans. The Chicago heat wave in 1995 and European 2003 and 2010 heat waves are all associated with blocking, but blocking can result in the persistence of any type of weather. Examples include the 2007 Texas floods, and cold winters in China in 2008, and in Europe 2005/2006 and 2010/2011.

5.3.3 What causes the seasons?

In Box 2.4 in Chapter 2 we learned of the importance of seasonality in driving health status, especially in rural populations in developing countries; but what causes the seasons in the first place? A very common misconception is that summer and winter are caused by changes in the distance from the Earth to the sun over the course of the year. The Earth is closest to the sun on about 3 January, i.e., in the Southern Hemisphere (*austral*) summer, and furthest from the sun on about 4 July, in the austral winter (Figure 5.10). The effect of the change in distance is small: the seasons in the Northern Hemisphere are slightly milder than they would be if the distance were the same all year, while those in the Southern Hemisphere are slightly more extreme. Instead, it is the dates of the solstices that are important in determining the seasons. The solstices are the dates on which either the most sunlight or the least sunlight is experienced; it is changes in the amount of sunshine during the year that cause the seasons.

The amount of sunshine received is much more strongly affected by the length of the day and how high in the sky the sun gets at high noon (§ 5.3.1). Winter is cold because the day is short and because the sun stays near the horizon. The

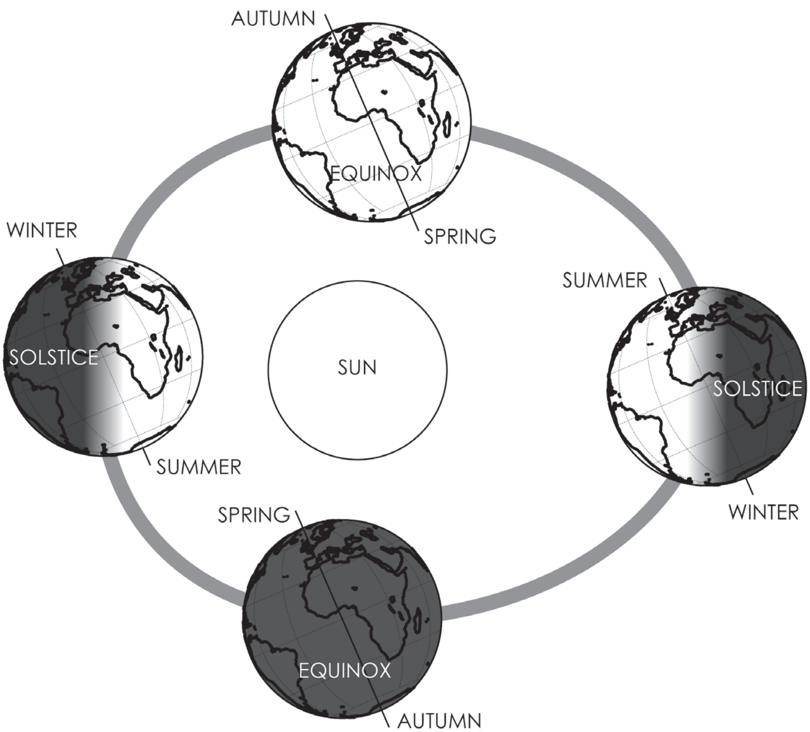

FIGURE 5.10 Illustration of the causes of Earth's seasons

changes in the length of day and the height of the sun in the sky are a result of the fact that Earth is tilted by about 23° – it does not spin around a vertical axis relative to its rotation around the sun. One way to understand that tilt is that the North Pole is not at the top of the Earth (Figure 5.10).

The tilt matters because its direction barely changes over the course of a human life-time, and so the axis is always pointing to the same star. Because Earth's tilt is practically constant, if you are standing in the right place you can use the Pole Star to find north, or the Southern Cross to find south, at any time of year. However, the constant tilt with respect to the stars also means that Earth's tilt is not constant with respect to the sun. In Figure 5.10, Earth's axis tilts toward the left. On about 20 June, the sun is also to the left of Earth, and so the North Pole is facing the sun, the sun is 23° above the horizon, and there is 24-hour daylight everywhere within 23° latitude of the North Pole (i.e., within the Arctic Circle, which is at about 67°N). At high noon, the sun is directly overhead at 23°N (the Tropic of Capricorn). Everywhere north of the Tropic of Capricorn the sun reaches its highest point above the horizon and it is the longest day of the year between the Tropic and the Arctic Circle. This date is the *boreal* (Northern Hemisphere) summer solstice (or the austral winter solstice from a Southern Hemisphere perspective).

On about 20 December, the sun is to the right of Earth, but Earth's axis is still tilted to the left, and so the North Pole is facing away from the sun. There is 24-hour night everywhere within the Arctic Circle, while everywhere else in the Northern Hemisphere experiences its shortest day on this date, and the sun's height above the horizon at high noon is the lowest for any time of the year. This date is the boreal winter solstice (or the austral summer solstice). At high noon the sun is directly overhead at 23°S (the Tropic of Cancer), and the Southern Hemisphere experiences its summer.

At the equinoxes (around 20 March and 22 September) the sun shines directly side-on to Earth – neither Pole points towards the sun. On these dates the sun is directly overhead at high noon at the equator, and everywhere on Earth has the same length of day and night. Those of us who live in the extratropics may use the position of the sun to identify north and south at high noon; within the Tropics the sun is sometimes to the north and sometimes to the south, depending on the time of year.

5.3.4 How do the seasons differ spatially?

As discussed in § 5.2.2, and illustrated in Figure 5.7, the seasons are most pronounced in high latitudes because of the strong contrast in the amount of sunlight between winter and summer at these latitudes. Within the tropics, the sun passes directly overhead twice per year. At the equator, these dates coincide with the equinoxes, but the length of day and the height of the sun in the sky does not change very much throughout the year. Therefore, 'summer' and 'winter' are not meaningful ways of defining the seasons here. Instead, close to the equator, the seasons are more meaningfully defined by rainfall than by temperature, although in some places very close to the equator even the rainfall remains nearly constant throughout the year.

Where the sun is directly overhead, Earth's surface is heated most intensely, and so, assuming there is a source of moisture, evaporation and convection (§ 4.2.2) are likely to be strong. There is a band of heavy rainfall that closely follows the sun as it moves north and south across the equator through the course of the year. This band of heavy rainfall is associated with the Inter-Tropical Convergence Zone (ITCZ). Strictly speaking, the ITCZ is defined in terms of winds rather than rainfall: it marks the line at which air from the Northern Hemisphere and the Southern Hemisphere converge into the low pressure that forms because of strong heating by the overhead sun. Nevertheless, the band of heavy rainfall closely follows the movement of the ITCZ, although in some longitudes (such as the south-eastern Pacific and the Middle East) there is too little moisture in the air for much rainfall to occur. The northern and southern limits of the ITCZ are modified by the positions of continents; it migrates furthest from the equator over land because the land heats up and cools down more quickly than the sea, and so it interacts with some of the main monsoon systems (§ 5.2.3). Between the northern and the southern limits, the ITCZ passes overhead twice per year and so many of these areas have two rainy seasons (as in much of East Africa, for example). Closer to the limits there is only the one rainy season (such as over continental South Asia and the Sahelian belt of West Africa).

Other areas with a single rainfall season include areas near the sub-tropics, such as the Mediterranean, Southern California, central Chile and the south-western parts of Australia and South Africa. These areas typically experience a wet winter and a hot dry summer because the subtropical deserts shift poleward in summer (§ 5.2.5) in response to the movement of the ITCZ. In the mid-latitudes, most areas have year-round rainfall, with large-scale rainfall dominating in the winter, and convective rainfall possibly becoming more important in the summer.

As well as geographical differences in the amplitudes of the seasons, there are also geographical differences in the lag in the seasons, and the lag can be strongly asymmetric. The lag tends to be longest in areas with coastal climates (§ 5.2.3), and is most extreme near the Poles. At the Poles, summer is delayed by only two or three weeks, but cooling continues throughout the winter until the sun starts to re-appear (March in the Northern Hemisphere, September in the Southern Hemisphere). As a result, near the North Pole, July is the warmest month, but occurs only five months after the coldest month, which is February. At the South Pole, because of the long distance from the sea, the lag is even more extreme: December is the warmest month, but occurs only four months after the coldest month, August.

5.3.5 How much does climate vary?

Just as latitude has a strong effect on changes in temperature throughout the year (§§ 5.2.3 and 5.3.4), it also dominates the variability in temperature from year-to-year (Figure 5.11 top). Temperature is most variable at high latitudes, partly because changes in wind direction can have a big influence on temperature especially in winter (§ 5.2.3), and partly because of sensitivity to changes in snow and ice cover (§ 4.2.7). Differences in temperature variability between inland and coastal areas are also evident: variability is stronger inland away from the dampening effect of the oceans (§ 5.2.3). Along the equator in the eastern and central Pacific Ocean temperature variability is relatively high for oceanic areas: the variability here is related to El Niño (Box 5.1).

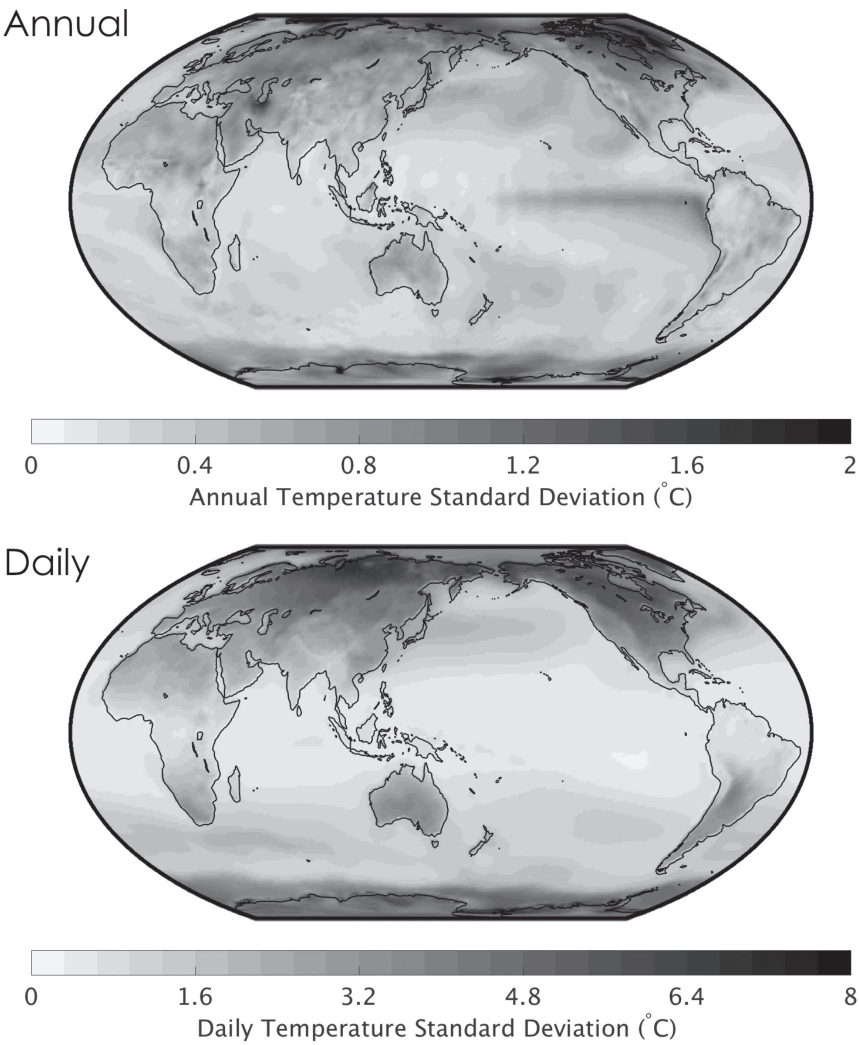

FIGURE 5.11 Year-to-year (top) and day-to-day (bottom) variability in temperature, as measured by the standard deviation, 1981–2010.
Data source: ECMWF Interim Reanalysis, for 1981–2010[1]

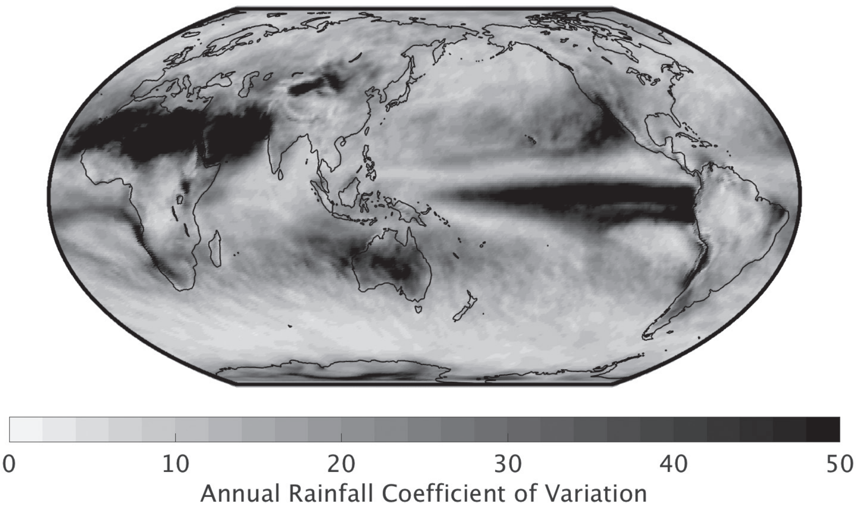

FIGURE 5.12 The coefficient of variation of annual rainfall, 1981–2010.
Data source: ECMWF Interim Reanalysis, for 1981–2010[1]

The spatial pattern of variability in temperature from day-to-day (Figure 5.11 bottom) is similar to that for variability from year-to-year (Figure 5.11 top), but temperatures vary from day-to-day at least four times more than they do from year-to-year over most of the globe outside of the tropics. Only over part of the tropical Pacific Ocean where El Niño occurs (Box 5.1) does temperature vary from year-to-year more than it does from day-to-day.

Comparing variability in rainfall at different locations is more complicated than for temperature. Because rainfall variability is affected by the average, the coefficient of variation (the standard deviation divided by the average) is used to control for this dependency (Figure 5.12). By this measure, variability is highest in the deserts, but generally is low in the mid-latitudes and humid tropics, and high in the subtropics.

Over much of the world, temperature and rainfall anomalies are related: if it is unusually wet or dry, it is frequently either unusually hot or cold at the same time. However, the nature of this relationship depends on location and season, and even time of day:

- Over the warm oceans, temperature increases with rainfall because of increased evaporation (i.e., temperature and rainfall are positively correlated).
- Over land, and especially in the tropics, daytime temperature decreases with increased rainfall because of increased cloudiness which blocks the sun's rays (i.e., temperature and rainfall are negatively correlated), but night-time temperature is positively correlated because of the blanketing effect of the clouds.
- Similarly, over the mid-latitude and subtropical land areas, summer daytime temperatures and rainfall are negatively correlated for the same reason.

- Near the poles, an increase in temperature in winter often occurs at the same time as an increase in snowfall because warmer air can hold more moisture (§ 4.2.3; i.e., temperature and rainfall are positively correlated).

All the correlations described above involve local relationships between rainfall and temperature, but they can be confounded if there are large-scale effects on the local climate, such as through El Niño.[9] During El Niño conditions, for example, maximum and minimum temperatures increase over much of the globe (Box 5.1), and where there is also an effect on rainfall the changes in temperature related to El Niño will be added to those associated with changes in rainfall. The resulting effect can become complicated.

5.4 Why does climate vary temporally?

Just as weather changes from day-to-day and week-to-week, so also does the climate change from month-to-month, from year-to-year, and beyond (Table 5.1). Some of this variability is as an effect of preceding weather and climate; these causes are understood as 'internal' to how climate works. Internal causes of variability include feedbacks or interactions in which the weather changes the Earth's surface, which in turn changes the weather. Asking which change comes first is a chicken and egg problem. Other causes of climate variability occur independently, perhaps because of human activity or because of major geophysical activity, or perhaps because of changes in how the Earth is heated by the sun. Independent changes are described as 'external'.

5.4.1 Internal causes of climate variability

Weather and climate are naturally variable, and primarily for the very reasons we have known about for thousands of years: 'the wind bloweth where it listeth' (*John 3:8*); 'the winds … scatter clouds and rain' (*Sūrat al Mur'salāt 3*). In other words, the weather behaves like that sometimes. When we ask a question such as 'why is it foggy this morning?' we are usually satisfied with an answer that might describe moist winds from a warm ocean blowing inland where clear skies overnight have made the land-surface very cold. Such a response explains the fog in terms of the weather patterns, and any explanation of those weather patterns might refer in turn to air pressure patterns, jet streams, etc. – i.e., to other weather patterns. Climate is variable, in part because weather is so variable – the chaotic (as distinct from 'random') variability of weather means that the weather never repeats itself, just as it is impossible to get a pinball to follow exactly the same path. In some years, fog happens frequently because sometimes there are frequent cold nights and warm moist winds. And that is all the explanation there is. Sometimes. Sometimes, fog happens frequently because that warm ocean is warmer than it normally is, or those cold clear nights are colder and more frequent than normal, and there may be an explanation for those anomalies. We have already seen how important the Earth's surface is as a source of heat, of moisture, and because of how much sunlight it reflects (§§ 5.2.3, 5.2.4 and 5.3.2). Changes

TABLE 5.1 Timescales of weather and climate variability and trends, their causes and sources of uncertainty

Timescale of variability	*Timeframe*	*Primary drivers*	*Dominant sources of uncertainty for prediction*
Weather	Hours to 2 weeks	Preceding weather patterns	Initial conditions in the atmosphere (butterfly effect)
Sub-seasonal	1–4 weeks	Some large-scale weather patterns	Chaotic variability; model uncertainty
Seasonal	1–12 months	Earth's axial tilt drives basic seasonality; sea-surface temperature anomalies (especially those associated with El Niño – Southern Oscillation [ENSO]) sometimes affect the evolution of a particular season, primarily in the tropics	Chaotic variability; model uncertainty; initial conditions in the ocean
Multi-annual	1–10 years	Sea-surface and deeper ocean temperatures (e.g., ENSO and Atlantic variability); volcanic activity	Chaotic variability; initial conditions in the ocean; model uncertainty; timing of eruptions
Multi-decadal	10–30 years	Ocean temperatures and circulation; anthropogenic forcing	Model uncertainty; initial conditions in the ocean
Century	> 30 years	Anthropogenic forcing; solar variability	Greenhouse gas emissions; model uncertainty

in Earth's surface can therefore have an important effect on climate, and these surface changes may themselves be a result of unusual and/or prolonged weather. Ways in which Earth's surface and the air can interact to cause climate variability include via sea-surface temperatures, land temperature and soil moisture, and snow and ice cover.

5.4.1.1 Variability in Earth's surface

Sea-surface temperature anomalies affect evaporation and heating or cooling of the overlying air, and the effect can last for weeks, months or even longer, because it takes so much energy to change the temperature of water (§ 4.2.1). The effect on climate is strongest in the tropics[10] because changes in the amount of water evaporated from the sea are much larger in hot climates than in cold (§ 5.2.4).

BOX 5.1 EL NIÑO – SOUTHERN OSCILLATION (ENSO)

What is the El Niño – Southern Oscillation?

After the the seasons, the El Niño – Southern Oscillation (ENSO) is the most important example of climate variability. Although it occurs in the Pacific Ocean region, it can affect climate over many other parts of the globe. The ENSO consists of changes in the ocean (El Niño) and changes in the atmosphere (Southern Oscillation). An *El Niño* state occurs when the sea surface across much of the eastern and central Equatorial Pacific Ocean becomes unusually hot (Figure 5.13, top), by as much as 3 °C or even more in the strongest events. It is also possible for this part of the ocean to become colder than normal. Such cooling conditions are called *La Niña* (Figure 5.13, bottom). These warming and cooling periods typically last about 9–12 months, commencing around April or May, and lasting through to about March, and recur about every three to ten years.

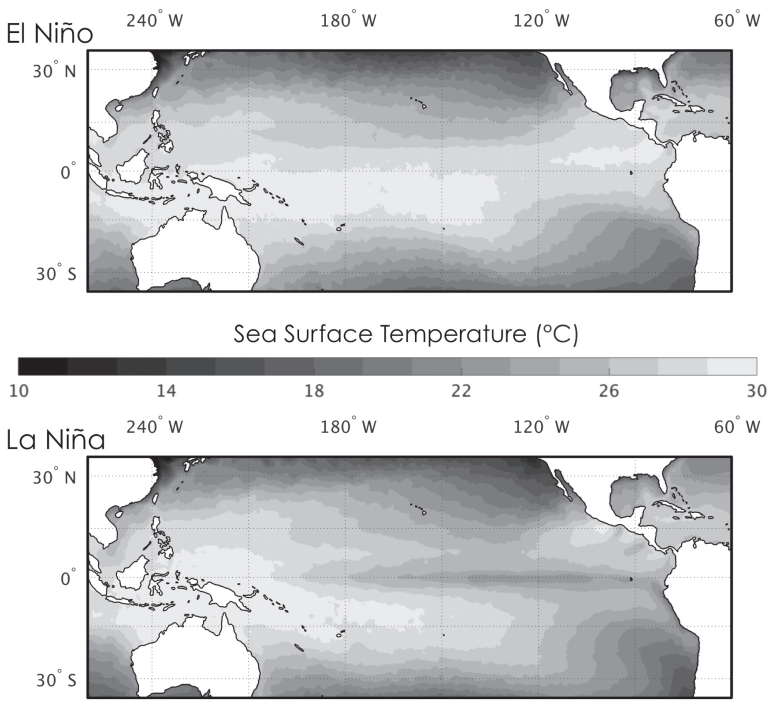

FIGURE 5.13 Sea-surface temperatures during (top) a strong El Niño event (Dec 1997–Feb 1998) and (bottom) a strong La Niña event (Dec 1998–Feb 1999). *Data source*: OISST2[11]

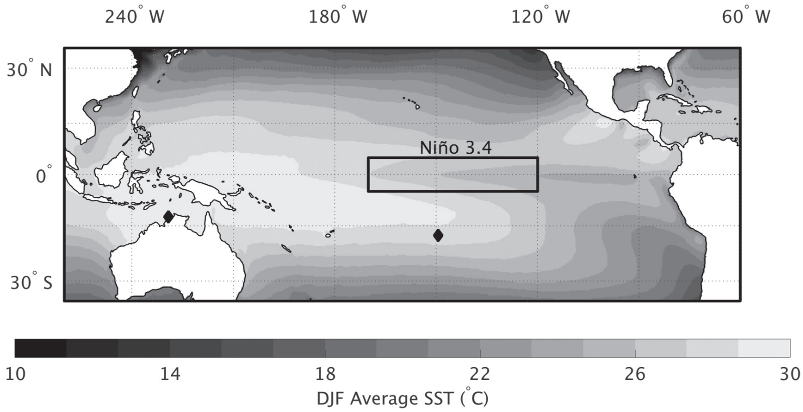

FIGURE 5.14 Average December–February sea-surface temperatures, 1982–2017. The black diamonds indicate the locations of Darwin and Tahiti, which are used in the calculation of the Southern Oscillation Index. *Data source*: OISST2[11]

The *Southern Oscillation* component of ENSO involves major changes in the position of high and low air pressure (§ 4.2.7) across the Pacific Ocean. The Oscillation measures air pressure differences between Darwin in Northern Australia, and at Tahiti in French Polynesia (at about 150°W in the central Pacific). In most years, air pressure is relatively low over Darwin, and high near Tahiti. This air pressure difference causes the Trade Winds, which blow from South America towards the western Pacific. (When describing the Pacific Ocean, it is easy to get confused about which side is east and which is west. The western Pacific is near Asia, which in most other contexts we understand as being the East; the eastern Pacific is near the Americas, which we understand as being the West.) Air pressure is low in the western Pacific because the sea is hot here (Figure 5.14), consistently exceeding about 28 °C, and the high humidity allows for heavy rainfall there (Figure 5.2). Further east, the cold ocean contributes to high atmospheric pressure and a drier climate.

The air and the sea are closely related across the equatorial Pacific in part because of an absence of the effects of land–sea contrasts, and so the Southern Oscillation and El Niño / La Niña vary in close relationship. During El Niño episodes, the warming in the eastern equatorial Pacific, where the sea is usually relatively cold (Figure 5.14), results in high temperatures extending across much of the ocean, thus weakening the air pressure difference. In strong El Niño conditions, sea-surface temperatures exceeding about 28 °C can extend across the entire width (Figure 5.13, top). During La Niña episodes, the cold eastern Pacific becomes even colder, and so the air pressure difference strengthens. La Niña thus looks much more like an enhancement of the average conditions (cf. Figures 5.12, bottom, and 5.13), whereas El Niño looks more like a change from the average (cf. Figures 5.13, top, and 5.14).

How often do El Niño and La Niña occur?

El Niño and La Niña conditions recur about every three to ten or more years, but their frequency and intensity vary inter-decadally[12] and inter-millennially. In the second half of the 19th and in the mid-20th century, for example, ENSO variability was relatively weak. The predictability of ENSO, and of climate variability in general, is poorer during these quiescent phases. Since the late-1960s ENSO has been more active, but we are unable to predict how long this phase will last.

Whenever El Niño and La Niña episodes occur, they typically dissipate in about March. However, occasionally, El Niño or La Niña can regenerate over the subsequent months, resulting in multi-year episodes. Since the mid-1800s, prolonged El Niño episodes have occurred every 20–60 years, the most recent occasions being in the early 1990s and mid-2010s. Prolonged La Niña episodes have occurred about twice as frequently, the latest being in the late 1990s.

How do El Niño and La Niña affect the climate?

El Niño and La Niña affect climate in a consistent way only in some parts of the world, and for only part of the year. The strongest effects are over the ocean, and only about 20% to 30% of land areas experience significant impacts on rainfall at least some part of the year. Different areas are affected in different seasons, and so rainfall over only about 15% to 25% of global land areas is affected in any particular season.[13] The strongest impacts are in the vicinity of the tropical Pacific Ocean, but rainfall in other parts of the globe can also be affected as large-scale climate patterns respond to a shift in the area of heavy rainfall between the eastern and central Pacific described above. The International Research Institute for Climate and Society has developed simple web-based tools that can be used to explore the relationship of ENSO to rainfall and temperature across the globe[i] or more specifically in Africa.[ii,14] Southern African rainfall is strongly impacted by ENSO with drought disasters associated with El Niño years[15] and high malaria incidence anomalies with La Niña years (see Case Study 5.2).

Impacts of El Niño and La Niña on air temperature are less well-studied than for rainfall. The warming of the equatorial Pacific Ocean during El Niño episodes does contribute to a notable increase in global average temperature and to warming over much of the tropics. The effects on air temperature are largely a combination of the increased heating from the warmer oceans, which extends into the Indian Ocean because of changes in wind, and of changes in cloudiness associated with shifts in rainfall patterns. Epidemics of malaria in highland regions of Ethiopia and Colombia have been notably associated with warmer El Niño years.[21]

CASE STUDY 5.2 IMPACT OF RAINFALL AND THE EL NIÑO – SOUTHERN OSCILLATION ON MALARIA IN BOTSWANA

Madeleine C. Thomson, IRI, Columbia University, New York, USA

Botswana has made substantial progress towards eliminating malaria over the last decade. Malaria cases reduced from 17,886 (0.97% prevalence) in 2008

TABLE 5.2 National de-trended (standardized) confirmed malaria cases (1982–2003) in Botswana during the malaria season (January–May)[17] and their relationship to December–February rainfall (estimated from Merged Analysis of Precipitation from the Climate Prediction Center [CMAP]). The table is ordered from low to high malaria incidence years. El Niño events (defined using the Oceanic Niño Index[iii]; see Box 6.3) in the months preceding and during the malaria season are mostly associated with low malaria anomalies while La Niña events are largely associated with high malaria years.[17,19] 1988 was affected by both El Niño and La Niña events

Year (ordered by malaria incidence anomaly – lowest to highest)	*Malaria incidence anomalies*	*Total DJF rainfall (in mm)*	*ENSO phase*
1992	–1.88	153	El Niño
1982	–1.57	162	
2002	–1.45	158	
1987	–1.18	170	El Niño
1983	–1.12	163	El Niño
2003	–1.11	200	El Niño
1995	–0.72	171	El Niño
1984	–0.62	174	
1991	–0.24	268	
1985	–0.14	222	La Niña
2001	–0.14	208	La Niña
1990	0.02	219	
1998	0.10	197	El Niño
1994	0.32	300	
1986	0.50	214	
2000	0.80	439	La Niña
1999	0.87	240	La Niña
1989	1.25	354	La Niña
1997	1.33	320	
1993	1.50	223	
1996	1.54	342	
1988	1.95	330	El Niño, La Niña

Abbreviations: DJF, December, January, February; ENSO, El Niño – Southern Oscillation.

to 311 (0.01% prevalence) cases in 2012.[16] Prior to recent investments in malaria control and elimination, year-to-year variability in malaria incidence was pronounced and could largely be explained by variations in rainfall during the December–February season once longer term trends (likely associated with control and treatment measures) had been removed.[17] Rainfall in much of Southern Africa closely follows sea-surface temperatures in the eastern equatorial Pacific, and major droughts or wet years are significantly associated with El Niño and La Niña (Box 5.1).[18] An analysis of malaria and climate 1982–2003 (Table 5.2) indicates the strength of the relationship between malaria, rainfall and sea-surface temperatures prior to the implementation of the elimination strategy. This relationship forms the basis for seasonal climate forecasts to provide early warning of malaria epidemics.[19] Incorporating climate into malaria impact assessment is necessary to avoid over or under-estimation of the impact of malaria interventions.[20] In March 2017 the Ministry of Health notified the public that the country was experiencing an unusually high level of malaria transmission following a period of heavy rains. Distinguishing between climate and other drivers (such as control factors) of year-to-year variations in malaria transmission continues to be of critical importance to malaria control managers.

Sometimes a change in sea-surface temperature is reinforced by changes in the weather that it causes. The air can change sea temperatures by various mechanisms: winds drive ocean currents, mix the warmer surface layers with colder subsurface water, and affect the rate of evaporation (the evaporation rate increases with strong winds), while evaporation and rain can have a direct effect on ocean temperature and an indirect one by affecting salinity. Salinity is important because it affects the density of water, which in turn drives ocean currents in a similar way to water temperature.[10] There are many examples of how the sea and the air affect each other to create changes in climate that operate on a wide-range of timescales. At some point, the interaction either breaks down or reverses, creating an 'oscillation' in which a climate anomaly develops, persists for a while, and reverses as a new mechanism takes over. The best known of these oscillations is the ENSO (Box 5.1). The ENSO is a Pacific Ocean phenomenon, but can affect climate around the world. The Atlantic Equatorial Mode and the Indian Ocean Dipole are analogous phenomena in the other tropical oceans, and which can have important regional impacts on climate.[10] Other examples of important climate oscillations are described in Box 5.2. Some of these oscillations can operate on timescales lasting many years.

Climate variability can also be caused by changes in the land surface by affecting how much sunlight is reflected (§ 4.2.7), especially in the presence of snow, and by affecting the amount of evaporation into, and heating of, the air (§§ 4.2.7 and 5.2.4). Examples include the influence of the Himalayan snow pack on the

BOX 5.2 CLIMATE OSCILLATIONS

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While ENSO is the most important mode of climate variations at the global scale and tends to have the greatest impact in the tropics, other important climate oscillations exist with more localized impacts.

North Atlantic Oscillation

The North Atlantic Oscillation (NAO) is a large-scale pattern of natural climate variability characterized by a seesaw difference in air pressure between the Azores and Iceland. The Oscillation has important effects on rainfall and temperatures across the eastern United States, and much of Europe, extending as far as the Middle East. Successfully predicting the Oscillation is key to making accurate seasonal forecasts of climate in these areas.[10] Until recently, models were unable to predict the Oscillation, but recent improvements in model resolution and initialization schemes seem to be enhancing the skill at seasonal timescales.

Madden-Julian Oscillation

The Madden-Julian Oscillation (MJO) involves an area of strong convection and heavy rainfall that moves eastward through the global tropics. The eastward movement occurs at varying speed, and can take from 30 to 60 days to circle all the way around the Earth. For areas in the tropics, the Oscillation can bring a series of wet and dry spells. Some effects outside of the tropics can also be felt. Recent advances in numerical modeling of the MJO are providing promise for forecasting weather at timescales of one week to one month, mostly because of improved modelling of rainfall processes.

South Asian monsoon, and the importance of soil moisture deficits as precursors to heatwaves.[10]

5.4.2 External causes of climate variability

Some changes to the Earth's surface occur independently, perhaps because of human activity or because of major geophysical activity. However, changes in Earth's surface are not the only external cause of climate variability: the amount of energy received from the sun can vary; and although Earth's surface provides heat and moisture to the air, how much of that heat and moisture the air retains can change because of the composition of the air. Such changes in the composition of the air can be natural or human-induced.

5.4.2.1 Volcanoes

The impact of a volcanic eruption on climate depends on much more than the violence of the eruption: the direction, location and chemistry of the eruptions are all important. Some volcanic eruptions, such as Mount St Helens in 1980, explode laterally, and so the dust and ash stay near Earth's surface where they can be washed out of the atmosphere by rain in a few days. Other eruptions, like Mount Pinatubo in 1991, explode vertically, and so the emissions can get high into the atmosphere where they can remain for months or even years, and cool the Earth by blocking sunlight from reaching the surface. These vertical eruptions have longer-lasting impacts than more lateral eruptions. For example, Mount Pinatubo's eruption cooled the Northern Hemisphere climate by about 0.6 °C as well as contributing to widespread decreases in rainfall in the tropics.[22] These impacts are considerably more than that of a major El Niño (Box 5.1). The very largest eruptions can have devastating impacts, and can change the climate for millennia if the cooling is sufficient to cause widespread snowfall and freezing.

Because of the direction of high altitude winds, particles from volcanic eruptions that do get above about 15 km are slowly transported towards the Poles (in the same way that ozone is transported towards the Poles; Box 4.4). Therefore, eruptions from volcanoes near the equator may have a more widespread impact than high latitude eruptions. The cooling effect is strongest when the eruption emits large volumes of sulphates, since these form aerosols that are highly effective in blocking sunlight. Indeed, sulphate aerosols from a series of relatively small near-equatorial eruptions are partly responsible for a slow-down in the global warming trend[23] over the first approximately 15 years of the 21st century. In addition to sulphates, volcanoes do emit CO_2, which contributes to the greenhouse effect (see § 9.3), but the average annual emission from volcanoes is less than 1% of human emissions, and so this greenhouse effect is negligible.

5.4.2.2 Solar variability

The amount of energy emitted by the sun varies on a fairly strict cycle of about 11 years, but the strength of these cycles itself changes on a less-predictable timeframe. One manifestation of these changes is the appearance of dark spots on the sun, which represent areas of stronger activity. There are reliable records of these sunspots extending back hundreds of years. During a quiescent period in the 1800s, decreased solar activity contributed to the development of the Little Ice Age, during which European winters were bitterly cold. Solar activity picked up in the mid-1800s, about the same time as the industrial revolution was causing an increase in greenhouse gas concentrations. Solar activity seems to have reached a peak in the 1990s, and has been decreasing since.

At timescales of thousands of years, changes in Earth's orbit affect how far the Earth is from the sun at different times of year, as well as by how much the amount of sunlight changes at different latitudes over the course of a year. These orbital changes can be projected forward and backwards in time many hundreds of thousands of years because they are based on the gravitational effect of the planets and the moon, whose movements are known in detail. The changes match exceptionally well with the advance and retreat of Ice Age conditions.

5.4.2.3 Atmospheric composition

The sun heats the Earth and the Earth heats the air. However, just as most of the air is transparent to most of the sun's radiation, so also the air is transparent to some of Earth's emitted radiation. We have already seen that ozone can absorb certain types of radiation that other gases cannot (§§ 4.2.6 and 4.2.7); in fact, each gas is able to absorb different types of radiation. Gases that absorb Earth's radiation are called greenhouse gases. On Earth, the main greenhouse gases are water vapour, carbon dioxide and methane. The most effective greenhouse gases are those that absorb radiation that other gases do not: similarly, a small board that can block a hole in the window will likely insulate your house more effectively than adding a second layer of loft insulation over tens of square metres. Because greenhouse gases can plug these metaphorical holes, they can be important even when their concentrations in the air are low. If Earth had no greenhouse gases it would be more than 30 °C colder than it is.

Changes in the amount of greenhouse gases in the air will affect its temperature, and thereby can alter the climate. Water vapour is easily the most abundant greenhouse gas on Earth, and its effects may even be sensible from day-to-day: cloudy nights are generally so much warmer than clear nights, for example (see further discussion in § 5.3.1). Because evaporation generally increases as temperatures rise, water vapour can serve to enhance the warming caused by other greenhouse gases. It can therefore be misleading to quote the warming potential caused by an increase in the concentration of a greenhouse gas in isolation. There are many complicating feedbacks.

After water vapour, carbon dioxide (CO_2) is the next most abundant greenhouse gas. It occurs naturally in the atmosphere, and natural variability is clearly evident in its annual cycle through the effects of plant growth. There is slightly over twice as much land in the Northern Hemisphere than in the Southern, and that difference is even greater when one considers only the latitudes with distinct growing seasons. As a result of this inequality in land distribution, concentrations vary by about 1.5% over a year, peaking at the end of winter before the Northern Hemisphere spring when plant growth starts to absorb the gas, and reaching a minimum at the end of summer when the leaves begin to fall. That 1.5% fluctuation is too small to have a significant effect on climate, but CO_2 concentration has increased by about 40% over about the last 250 years because of human activities, primarily through the burning of fossil fuels. The current rate of increase over about three years is equivalent to the increase between autumn and spring as a result of the annual cycle mentioned above.

Methane is another important greenhouse gas, and has increased by about 150% over the same period as CO_2. The methane increase is primarily a result of livestock farming. There is concern that melting of the permafrost in high latitudes because of global warming will release large amounts of additional methane into the air, thus enhancing the greenhouse effect further.

5.5 Conclusions

Spatial and temporal variations of temperature are much simpler than those of rainfall at virtually all scales. It is important to understand these scales of variability in space and time to obtain some idea of the necessary resolution of data for any analyses of climate–health relationships. Such an understanding can also contribute to an awareness of some of the limitations of these analyses given the constraints that data availability may impose. The following chapter provides an introduction to the nature, availability and limitations of climate data.

Notes

i https://iridl.ldeo.columbia.edu/maproom/ENSO/Impacts.html.
ii http://iridl.ldeo.columbia.edu/maproom/Health/Regional/Africa/Malaria/ENSO_Prob/ENSO_Prob_Precip.html.
iii http://origin.cpc.ncep.noaa.gov/products/analysis_monitoring/ensostuff/ONI_v5.php.

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